Soil Water Balance and Ecosystem Response to Climate Change

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ABSTRACT: Some essential features of the terrestrial hydrologic cycle and ecosystem response are singled out by confronting empirical observations of the soil water balance of different ecosystems with the results of a stochastic model of soil moisture dynamics. The simplified framework analytically describes how hydroclimatic variability (especially the frequency and amount of rainfall events) concurs with soil and plant characteristics in producing the soil moisture dynamics that in turn impact vegetation conditions. The results of the model extend and help interpret the classical curve of Budyko, which relates evapotranspiration losses to a dryness index, describing the partitioning of precipitation into evapotranspiration, runoff, and deep infiltration. They also provide a general classification of soil water balance of the world ecosystems based on two governing dimensionless groups summarizing the climate, soil, and vegetation conditions. The subsequent analysis of the links among soil moisture dynamics, plant water stress, and carbon assimilation offers an interpretation of recent manipulative field experiments on ecosystem response to shifts in the rainfall regime, showing that plant carbon assimilation crucially depends not only on the total rainfall during the growing season but also on the intermittency and magnitude of the rainfall events.

Keywords: soil moisture, soil water balance, plant water stress, stochastic processes, ecosystem response, climate change.

The terrestrial hydrologic cycle is an example of a manifold system whose understanding requires a synergistic use of empirical observations, simple and detailed models, and theoretical and numerical analyses. The importance of its appropriate description is evident, and so are its implications for water resource availability, flood occurrence, biogeochemistry, and plant conditions (e.g., Noy Meir 1973; Stephenson 1990; Easterling et al. 2000; Allen and Ingram 2002; Milly et al. 2002). Changes in rainfall regime and hydrologic cycle due to increased concentration of greenhouse gases have already been detected and are predicted to further increase (e.g., Easterling et al. 2000), and it is thus crucial for the scientific community to concentrate its efforts on an improved understanding and prediction of the ecological responses to such changes, employing proper combinations of experiments and theoretical analyses to overcome the inherent difficulty of dealing with a complex nonlinear systems with essential stochastic components (Clark et al. 2001).

There is growing evidence that the predicted changes in rainfall regime due to climate change will reduce ecosystem net primary productivity and possibly induce shifts in community composition (Knapp et al. 2002). Plant productivity and water stress as well as soil biogeochemistry are strongly controlled by the pulsing and unpredictable nature of soil moisture dynamics. Therefore, accounting only for changes in mean responses to climatic variability is not sufficient for a realistic investigation of impact of climate change on ecosystems, which instead must account for the stochastic component of the hydrologic forcing and its possible alterations in terms of frequency and amount of rainfall events. Such alterations are responsible for modifying soil moisture dynamics and the temporal structure (i.e., intensity, duration, and frequency) of periods of water stress and impaired plant assimilation (Porporato et al. 2001). In fact, soil moisture deficit induces a reduction of plant water potential that, in turn, may cause dehydration, turgor loss, xylem cavitation, stomatal closure, and reduction of photosynthesis (e.g., Nilsen and Orcutt 1998). Even maintaining the same total rainfall, an increase in the intensity of rainfall events, concomitant with a reduction in their frequency, will affect soil moisture dynamics and plant conditions in a manner that depends on the soil and plant physiological characteristics at the site.

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From the modeling viewpoint, the very large number of processes that make up the dynamics of the soil water balance and the extremely large degree of nonlinearity and space-time variability of hydrological and ecological phenomena (Clark et al. 2001; Porporato and Rodriguez-Iturbe 2002) call for simplifying assumptions at different levels. Whenever possible, the development of a low-dimensional description in which the dominating deterministic (and possibly nonlinear) components are separated from the high-dimensional (i.e., stochastic) environmental forcing is especially valuable. In particular, simple models of soil moisture dynamics have been used to capture the essential features of the terrestrial water cycle and the resulting vegetation response (Eagleson 1978; Milly 1993; Rodriguez-Iturbe et al. 1999; Laio et al. 2001). From the resulting analytical solutions, the role of the controlling parameters clearly emerges, offering a theoretical framework whose generality surpasses that of more complicated models that require cumbersome numerical simulations.

Here we follow a minimalistic approach to the modeling of the soil-plant-atmosphere system by further simplifying a previous stochastic model of soil moisture dynamics (Rodriguez-Iturbe et al. 1999; Laio et al. 2001) and coupling it with a simple representation of the nonlinear link between carbon assimilation and soil moisture at the daily timescale. Our aim is to offer a very parsimonious yet realistic representation of soil water balance that captures its essential components: the water-holding capacity of the soil, which is a function of soil and root characteristics and is responsible for the threshold-like nonlinearity that triggers deep infiltration and surface runoff; the soil-moisture dependence of evapotranspiration and photosynthesis; and the intermittency and unpredictability of rainfall, whose variability in terms of both frequency and depth of events is crucial not only for the soil water balance but also for the ecological processes (Noy Meir 1973; Rodriguez-Iturbe et al. 1999).

Methods

A Simple Stochastic Model for Soil Moisture Dynamics

We interpret the soil moisture dynamics at the daily timescale, treating the soil as a reservoir with an effective storage capacity that is intermittently filled by rainfall events in the form of pulses of random depth. Soil water losses occur via evapotranspiration, deep infiltration, and surface runoff (Milly 1993; Rodriguez-Iturbe et al. 1999). Both the vertical and horizontal spatial variability are neglected, assuming that the propagation of the wetting front and the soil moisture redistribution over the rooting zone are negligible at the daily timescale. Lateral water flow is also neglected, thus excluding regions with marked topographic effects.

We consider the relative soil moisture s (dimensionless), vertically averaged over the rooting zone of depth Z_r (cm), as the state variable describing the dynamics of the soil water balance. Accordingly, the total volume of soil water per unit ground area at a given time t is snZ_r (cm), where n is the vertically averaged soil porosity (volume of voids/total volume of soil, i.e., dimensionless). Both n and Z_r are assumed to be time-invariant parameters. To model the soil water balance dynamics we assume that when s exceeds a given threshold s_1, the rainfall in excess of the available storage capacity is immediately lost by runoff and deep percolation or drainage, LQ (cm day^{-1}). The empirical parameter s_1 depends on the type of soil and is typically comprised between the so-called field capacity (i.e., the soil moisture level where drainage becomes negligible) and complete saturation ($s = 1$). Notice that a parameter similar to s_1 was also adopted in previous studies of the soil water balance (Milly 1993; Federer et al. 2003). As discussed in Laio et al. (2001), the present approach to infiltration modeling is useful when the Dunne or saturation-from-below mechanism of runoff formation is dominant compared to the Hortonian runoff (i.e., rainfall intensity exceeding the soil saturated hydraulic conductivity); this is often the case for vegetated surfaces with negligible topography and absence of soil crusting.

Evapotranspiration, ET (cm day^{-1}), is assumed to decrease linearly from a maximum value (sometimes referred to as potential evapotraspiration), ET_{max}, under well-watered conditions ($s = s_1$) to 0 at the wilting point ($s = s_w$). The reduction of evapotranspiration with decreasing soil moisture is a well-established fact that can be ascribed to increased resistances to soil water transport within the soil-plant-atmosphere continuum when soil water potential is reduced (e.g., increased soil-root resistance, progressive cavitation in the xylem conduits, stomatal closure). At a point scale in space, a marked nonlinearity is typically present in the evapotranspiration–soil moisture relationship (Laio et al. 2001; Daly et al. 2004a); however, at larger scales (e.g., regional), the temporal variability and special heterogeneity of hydrological processes tends to significantly broaden the linear rise in the evapotranspiration–soil moisture relationship (Wetzel and Chang 1987, 1988; Crow and Wood 2002). A similar tendency to linearizing the soil water losses was also noticed in a theoretical analysis of the mean soil moisture dynamics (Laio et al. 2002).

Rainfall input, $R(t)$ (cm day^{-1}), is modeled as a marked Poisson process with frequency λ (day^{-1}) and events carrying a random depth of rainfall with exponential distribution of mean α (cm). Such a model has been shown to provide a simple and realistic representation of rainfall at

the daily timescale for different hydroclimatic regimes (Milly 1993; Rodriguez-Iturbe et al. 1999). It is particularly useful to explicitly and efficiently account not only for changes in mean rainfall rates but also for changes in frequency and amount of rainfall events. Thus for a typical growing season of duration T_{seas}, the total average rainfall amount is $\alpha\lambda T_{seas}$.

According to the modeling scheme describe above, the soil moisture balance equation can thus be written as

$$nZ_r\frac{ds}{dt} = R(t) - \text{ET}[s(t)] - \text{LQ}[s(t), t]. \quad (1)$$

Because of the forcing term $R(t)$, equation (1) is a stochastic differential equation that requires a solution in probabilistic terms (see "Normalization and Probabilistic Steady State Solution"). Details about the implications of some of the modeling assumptions and their possible generalization can be found in Laio et al. (2001*a*). Applications to natural ecosystems of this more complete model and its possible extensions to cases where seasonal trends in rainfall and evapotranspiration are important can be found in Laio et al. (2001*b*, 2002) and Porporato et al. (2003). A critical discussion of the implications of neglecting the soil moisture vertical distribution (i.e., propagation of the wetting front, hydraulic lift, etc.) at the daily timescale can be found in the literature (Laio et al. 2001*a*; Guswa et al. 2002; Federer et al. 2003). We only note here that the errors introduced with such a simplification tend to be negligible compared to the uncertainties in the external hydroclimatic forcing.

Normalization and Probabilistic Steady State Solution

Defining $x = (s - s_w)/(s_1 - s_w)$ as the "effective" relative soil moisture and $w_0 = (s_1 - s_w)nZ_r$ as the maximum soil water storage available to plants, the governing quantities of the process are w_0, α, λ, and ET_{max}. According to dimensional analysis, these quantities can be grouped into two dimensionless numbers as $\gamma = w_0/\alpha$ and $\lambda/\eta = (\lambda w_0)/\text{ET}_{max}$ or $\gamma = w_0/\alpha$ and $D_I = (\gamma\eta)/\lambda = \text{ET}_{max}/\langle R \rangle$, where D_I is Budyko's dryness index, η is the normalized evapotranspiration loss under well-watered conditions, ET_{max}/w_0, and $\langle R \rangle$ is the mean rainfall rate, $\langle R \rangle = \alpha\lambda$ (Milly 2001). Physically, this means that the terrestrial water balance is governed by the ratio between the soil storage capacity and the mean rainfall input per event and either the dryness index, that is, the ratio between the maximum evapotranspiration and the mean rainfall rate, or λ/η, which is the ratio between the rate of occurrence of rainfall events and the maximum evapotranspiration rate. Such dimensionless groups define the interaction of the most

important climate, soil, and vegetation parameters in controlling soil moisture dynamics.

Following Rodriguez-Iturbe et al. (1999), the master equation of the probability density function (PDF) of x can be obtained and solved analytically for steady-state conditions. The result is a truncated gamma distribution; that is,

$$p(x) = \frac{N}{\eta}x^{(\lambda/\eta)-1}e^{-\gamma x} \quad (2)$$

for $0 < x \le 1$. The normalization constant is N, and

$$N = \frac{\eta\gamma^{\lambda/\eta}}{\Gamma(\lambda/\eta) - \Gamma(\lambda/\eta, \gamma)}, \quad (3)$$

where $\Gamma(\cdot)$ is the gamma function and $\Gamma(\cdot, \cdot)$ is the incomplete gamma function (Abramowitz and Stegun 1964). Because of the truncation, the shape of the soil moisture PDF depends on the scale parameter γ, as well as the shape parameter λ/η.

The mean effective relative soil moisture, $\langle x \rangle$, can also be obtained analytically,

$$\langle x \rangle = \frac{1}{\eta\gamma}(\lambda - Ne^{-\gamma}), \quad (4)$$

and so can the normalized water balance,

$$1 = \frac{\langle \text{ET} \rangle}{\langle R \rangle} + \frac{\langle \text{LQ} \rangle}{\langle R \rangle} = D_I\langle x \rangle + \frac{\langle \text{LQ} \rangle}{\langle R \rangle}, \quad (5)$$

where $\langle \text{ET} \rangle = \langle x \rangle \text{ET}_{max}$. Equation (5), with the aid of equation (4), describes the partitioning of the rainfall input into evapotranspiration and deep infiltration plus runoff as a function of the governing parameters of the climate, soil, and vegetation system.

Results

The model proposed here is a minimalist representation of soil moisture dynamics. In the following applications we show that it provides a realistic description of the terrestrial water balance under a wide range of conditions. Moreover, when suitably combined with a threshold defining incipient plant water stress, it also offers a useful framework to link hydrologic fluctuations to vegetation response.

Budyko's Hydroclimatological Description

A well-known hydroclimatological relationship developed by Budyko (1974) describes the average terrestrial water balance by means of a semiempirical curve (fig. 1, *dots*), which represents the fraction of the rainfall that is evapotranspired as a nonlinear function of the dryness index. Such a curve was tested on several river basins with different characteristics and synthesizes the average partitioning of the rainfall input into evapotranspiration and runoff plus drainage. The theoretical solutions of the water balance (eq. [5]) are also shown in figure 1 for different values of the governing parameter γ that contains the average rainfall depth and the soil water holding capacity (through the plant rooting depth and soil texture). Remarkably, for γ near 5.5, the model reproduces Budyko's curve very well. This means that using typical values of the parameters (e.g., average rainfall depth per event $\alpha = 1.5$ cm, relative soil moisture at the wilting point $s_w = 0.2$, relative soil moisture threshold for deep infiltration and runoff $s_1 = 0.85$, and porosity $n = 0.4$), Budyko's curve corresponds to a soil depth of approximately 30–35 cm. Such a value represents the average soil depth that is active from a hydrologic point of view. Interestingly, it also provides a good estimate of a typical effective rooting depth, which is in agreement with recent root surveys in water-limited ecosystems (Jackson et al. 2000; Schenk and Jackson 2002). Figure 1 also shows that, all the other parameters remaining the same, a tendency to have deeper roots implies moving upward in the diagram and therefore having an increase in evapotranspiration. However, this would not necessarily mean an improvement in plant conditions because it would also imply a change in the temporal structure (frequency, duration, and intensity) of water stress (Porporato et al. 2001).

Figure 1 also helps explain the effects of possible climate changes on Budyko's curve. As an example, depending on the degree to which evapotranspiration, rainfall regime, and plant characteristics are affected by climate change, alterations in the mean depth of rainfall per event will imply a vertical shift in the diagram while a shift along the *X*-axis will entail changes in potential transpiration and mean rainfall rate.

Classification of Soil Water Balances

The response of the soil water balance to the forcing by the climate-soil-vegetation system is synthesized by the PDF of the effective relative soil moisture. In particular, the behavior of the PDF as a function of the governing parameters (eq. [2]) provides a general classification of soil moisture regimes. Figure 2 shows that the PDF has qualitatively different shapes according to the values of the governing parameters. In the terminology of statistical mechanics, these changes may be interpreted as noise-induced transitions of the physical system (Horsthemke and Lefever 1984). The boundaries indicated in figure 2 may thus be used to define different hydroclimatic regimes: an "arid" regime (PDFs with 0 mode, i.e., a the wilting point), an
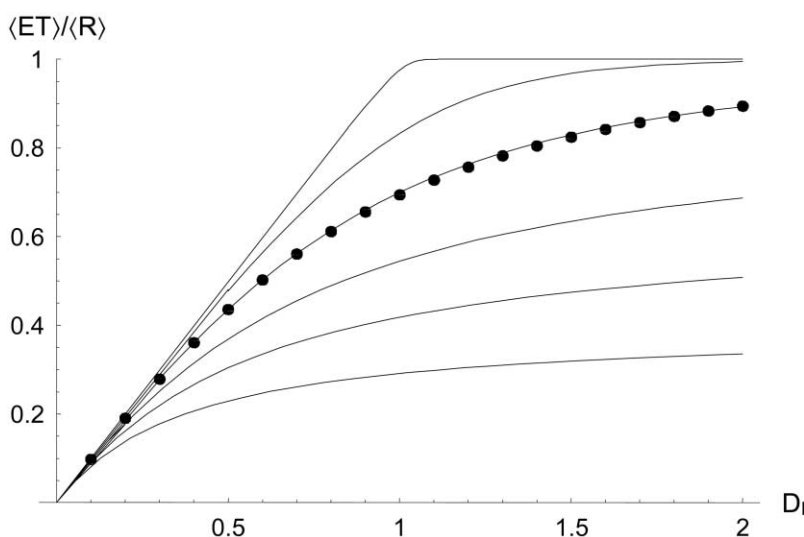

Figure 1: Fraction of total rainfall lost by evapotranspiration as a function of Budyko's dryness index for different values of the parameter γ. The dots represent the semiempirical curve of Budyko, $\langle ET \rangle / \langle R \rangle = \{D_I [1 - \exp(-D_I)] \tan h(1/D_I)\}^{0.5}$. The continuous line underlying the dots corresponds to $\gamma = 5.5$. As explained in the text, this refers to an average effective rooting depth of approximately 35 cm. From the lowest to the highest, the continuous curves refer to $\gamma = 0.5, 1, 2, 5.5, 20,$ and $1,000$, respectively.
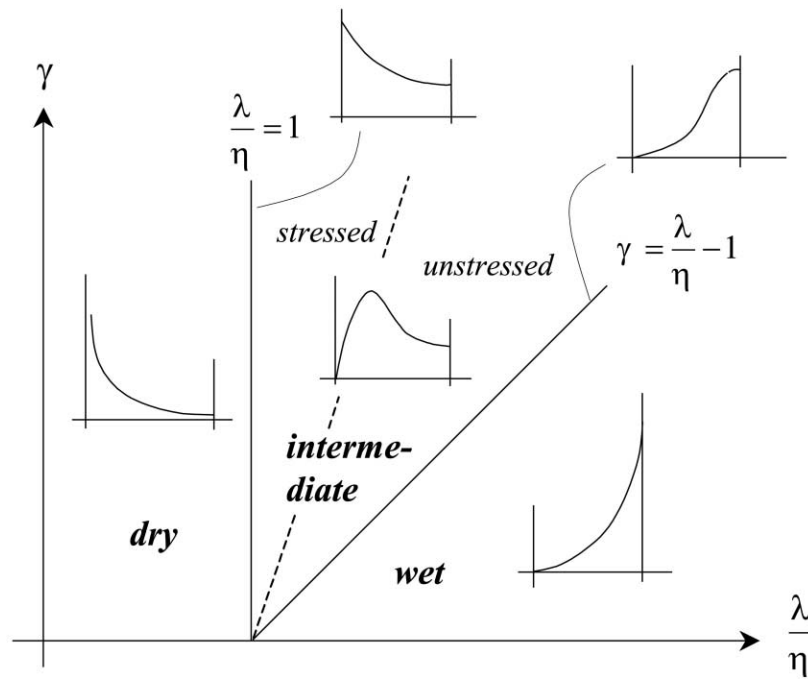
Figure 2: Classification of soil water balance, based on the shape of the soil moisture probability density function (PDF), as a function of the two governing parameters, λ/η and γ, that synthesize the role of climate, soil, and vegetation (see "Methods" for details). The dashed line, $\gamma = (1/x^*)[(\lambda/\eta) - 1]$, is the locus of points where the mode of the soil moisture PDF is equal to the threshold x^*, which marks the onset of plant water stress.

"intermediate" regime (corresponding to soil moisture PDFs with a central maximum), and a "wet" regime (with the mode at $x = 1$, i.e., well-watered conditions). A further distinction within the intermediate regime can be made on the basis of plant response to soil moisture dynamics. Defining x^* as a threshold (typically on the order of 0.3–0.4) marking the onset of plant water stress (Larcher 1995; Nilsen and Orcutt 1998; Porporato et al. 2001; Sperry et al. 2002), the dashed line of slope $1/x^*$ in figure 2 becomes the place where the mode of the effective relative soil moisture PDF is equal to x^* and thus where plants are more likely to be at the boundary between stressed and unstressed conditions. Accordingly, such a line may be used to divide water-stressed (or semiarid) types of water balance on the left side from unstressed ones on the right side.

Vegetation Response to Changes in Frequency and Amount of Rainfall

Climate change is presumed to impact the rainfall regime, especially in terms of frequency and intensity of rainfall events. It is therefore of extreme interest to be able to predict the effects of such changes on the frequency and duration of plant water stress and, in turn, on ecosystem productivity. When water stress appears, the daily leaf carbon assimilation rate (A) is reduced from its maximum value ($A_{\max}$) typical of well-watered conditions (Larcher 1995; Bonan 2002). Assuming that A is equal to $A_{\max}$ for $x > x^*$ and that it linearly decreases to 0 as soil moisture approaches the wilting point, the mean carbon assimilation rate during a growing season may be derived analytically as a function of climate, soil, and vegetation characteristics using a derived distribution approach from the probabilistic solution of soil moisture dynamics. While the analytical details are reported elsewhere (Daly et al. 2004*b*), here the results are used to analyze the recent findings of a 4-year manipulative experiment (Knapp et al. 2002; Fay et al. 2003) in which the ecosystem response of a native grassland to increased rainfall variability was investigated by artificially reducing storm frequency and increasing rainfall quantity per storm while keeping the total annual rainfall unchanged.

Figure 3 shows a comparison of the experimental results (Knapp et al. 2002) with the theoretical mean carbon assimilation as a function of the frequency of rainfall events for fixed total rainfall during a growing season. The ~20% decrease in measured net assimilation for the altered rain-
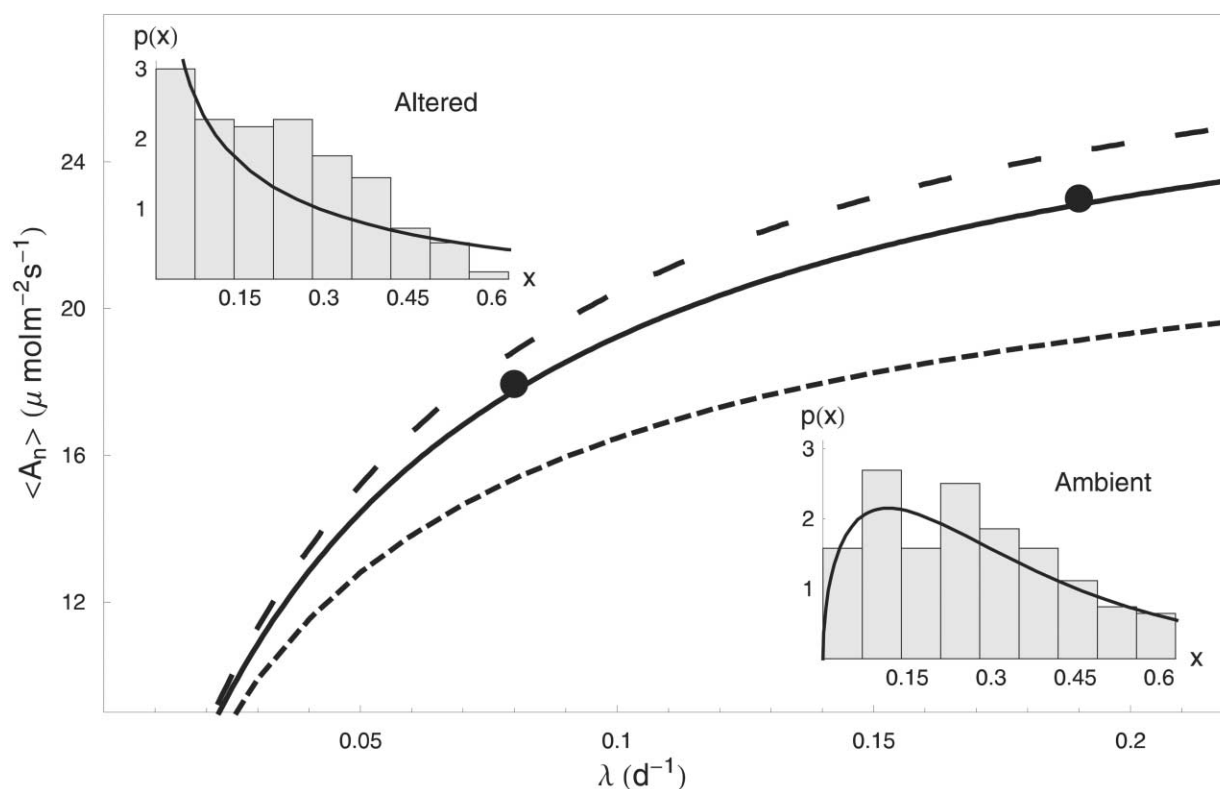
Figure 3: Mean daily carbon assimilation rate as a function of the frequency of rainfall events for constant total amount of precipitation during a growing season. The lines are the theoretical curves derived from the soil moisture probability density function, while the two points are field data published by Knapp et al. (2002), who compared the response of a mesic grassland to ambient rainfall pattern versus an artificially increased rainfall variability. The point on the right corresponds to the ambient conditions, and the point on the left corresponds to artificially modified conditions while keeping the total rainfall the same. The continuous line is for mean total rainfall during a growing season of 507 mm, the dashed line for 600 mm, and the dotted line for 400 mm. The two insets show observed and theoretical soil moisture probability density functions for ambient and altered conditions. The parameters used are $\lambda_{ambient} = 0.19$ day^{-1}, $\lambda_{altered} = 0.08$ day^{-1}, $n = 0.55$, $Z_r = 30$ cm, $E_{max} = 0.63$ cm day^{-1}, $A_{max} = 39$ μmol m^{-2} s^{-1}, $s_w = 0.12$, $s^* = 0.30$, and $s_1 = 0.8$.

fall pattern is well reproduced, going from a mean net assimilation of 23 μmol m^{-2} s^{-1} in natural condition to ~18.4 μmol m^{-2} s^{-1} when total rainfall was the same but concentrated in fewer events. As shown by the effective relative soil moisture PDFs, the dramatic shift in the rainfall frequency changes the grassland water balance from an intermediate to a dry one (cf. fig. 2 and fig. 3). The analysis also shows that in such a grassland ecosystem (Knapp et al. 2002), the impact on carbon assimilation of a decrease in total rainfall is more pronounced when such a decrease is accompanied by a reduction in the frequency of rainfall events. If the mean total rainfall is kept constant, then the sensitivity of mean assimilation to the frequency of rainfall events becomes much more pronounced for dry periods.

Conclusions

We have shown how the essential traits of the terrestrial water balance can be described by a simple stochastic model that explicitly accounts for the rainfall unpredictability both in terms of frequency and amount of rainfall events. A simple threshold to separate stressed and unstressed conditions gives a first-order representation of the plant nonlinear response to soil moisture dynamics. In this manner, it is possible to explain the main features of the soil water balance and the resulting ecosystem conditions under present-day and projected climatic scenarios. These results may be useful to generalize and better understand the results of field experiments as well as of more elaborated numerical models.

The present framework is also expected to be useful to

investigate the impact of different plant physiological characteristics (e.g., rooting depth, transpiration, and assimilation sensitivity to water stress) on the soil water balance at a site or to link plants' adaptation strategies to soil and hydroclimatic conditions. Possible extensions of the model to include seasonal components in rainfall and transpiration as well as to account for transient soil moisture dynamics at the start of the growing season due to the winter soil water recharge or spring snow melt could be devised to analyze their interaction with the stochastic hydrologic forcing during the growing season. Preliminary examples along these lines can be found elsewhere (Rodriguez-Iturbe et al. 2001; Laio et al. 2002).

Acknowledgments

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